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LAWS GOVERNING SURFACE PHYSICO-GEOGRAPHICAL PROCESSES

[figures in the appendix]

M. I. Budyko

One of the most important generalizations of modern physical ge-
ography is the conception of the physico-geographical process elaborated
in the important works of A. A. Grigor'yev ([5], [6], [7], and others).
These investigations have for the first time studied the interaction of
the varied physical, chemical and biological processes taking place in
the atmosphere, hydrosphere and lithosphere, from the unified viewpoint
of the totality of the physico-geographical phenomena.

It was established by A. A. Grigor'yev that in the investigation
of the physico-geographical process over periods of time, which are short
as compared with the duration of geological epochs, it is sufficient
to restrict our studies to "surface" physico-geographical processes,
which are determined by the basic four factors — climate, hydrology,
soil, and biology (first of all, the phytogeographical). Of decisive
importance in the development (intensity) of surface physico-geographical
processes, according to Grigor'yev, is the climatic factor, which de-
termines the development of the remaining hydrological, soil, and bio-
geographical processes.

The purpose of this work is the determination, from physical
considerations, of those quantitative characteristics of the climatic
factor, which, according to Grigor'yev's conception, determines: (1) the
zonal distribution of the hydrological areas, vegetation and soil for-
mations; (2) the intensity of the surface physico-geographical processes.
Determination of these characteristics permits also the enunciation of
a new principle of classification of climatic zones.

In order to solve the subject problem, this study utilizes the
data of some of the former works of this writer, particularly that pub-
lished in reference article [3].

1. Conditions of the thermal and aqueous regimes, which deter-
mine the climatic factor of the surface physico-geographical process,
can be characterized quantitatively by the equations of the thermal and
aqueous balances. These equations can be set up either for certain
layers, or for certain levels of atmosphere, hydrosphere or lithosphere.
However, in choosing certain levels (layers), we shall obtain materially
different relationships between the terms of the thermal and aqueous
balances. In order to find the equations of the thermal and moisture ex-
changes suitable for the analysis of the laws of the surface physico-
geographical process, it is necessary to consider the following.

The interaction of the climatic, hydrological, soil and biologi-
cal factors of the surface physico-geographical processes manifests it-
self in various ways throughout the entire thickness of the vast layer
encompassing the upper reaches of lithosphere and a considerable portion
of hydrosphere and atmosphere. However, the intensity of this interac-
tion changes noticeably with height. It is possible, in particular, to
note that, since the pertinent processes of soil formation, flow forma-
tion and animal-vegetable life on earth, take place either at the under-
lying surface, or in its immediate vicinity, the immediate effect of the
climatic factor on the remaining three factors must manifest itself pri-
marily at levels close to the level of the underlying surface. Therefore,
the development of the soil, hydrological and biogeographical processes
is effected mainly by the climatic characteristics of the thin layer ad-
jacent to the underlying surface.

Taking this into account, we shall take, as the characteristics
of the thermal and aqueous regimes of the climatic factor of the surface
physico-geographical process, the equations of the thermal and moisture
balances, set up for the level of the underlying surface. As is known
for mean-annual conditions over a period of years these equations are of
the following form (refer to $\boxed{3}$)

$$R = LE + P \qquad (1)$$

$$r = E + f \qquad (2)$$

where R is the radiation balance, E - evaporation, P - the turbulent thermal exchange between the underlying surface and the atmosphere, r - precipitation, f - drainage, L - the latent heat of vaporization. To the equations (1) and (2) we must add one semi-empirical equation connecting the terms of the thermal and aqueous balances (in future this supplementary equation will be referred to as the "connecting equation").

The "connecting equation" is derived as follows:

Before everything else, we note that in very dry soil all of the water received in the form of precipitation is retained by the molecular forces in the soil particles, and in the final analysis, is lost by evaporation. At these conditions (observed in deserts) the coefficient of drainage approaches zero.

Considering that the average aridity of soil increases with the increase of heat input by radiation and the decrease in precipitation, we establish that:

$$\frac{f}{r} \to 0 \text{ or } \frac{E}{r} \to 1 \text{ at } \frac{R}{Lr} \to \infty \qquad (3)$$

As $\frac{R}{Lr}$ decreases, $\frac{E}{r}$ will also decrease (there takes place some drainage), and with sufficiently high precipitation coupled with the correspondingly small input of heat by radiation, R, there will be reached a condition of constant overhumidification of the top soil layer. To assess the extent of evaporation in this case, it is necessary to consider that, in view of the unilateral nature of the vertical turbulent thermal conductivity, ("valve effect," refer to the study by this author and M. I. Yudin [4]) the sum total of heat flow from the atmosphere toward the earth, as a rule, is always small compared with the flow of heat from the earth into the atmosphere. Therefore, in all cases the

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inequality $LE < R$ holds. On the other hand, it is obvious that with small input of heat by radiation and for considerable precipitation, which insures relatively large evaporation, the surface of the soil cannot become materially overheated as compared with the air, in view of which the turbulent thermal loss from earth to the atmosphere becomes very small. Therefore, we can say that

$$LE \rightarrow R \quad \text{at} \quad \frac{R}{Lr} \rightarrow 0 \tag{4}$$

Conditions (3) and (4) determine the form of the function ϕ :

$$\frac{E}{r} = \phi \left(\frac{R}{Lr} \right) \tag{5}$$

for $\frac{R}{Lr} \rightarrow 0$ and for $\frac{R}{Lr} \rightarrow \infty$.

It must be stated that on the basis of analysis of the empirical data in accordance with precipitation and drainage, Schreiber (12) and Ol'dekop (10) have long since directed attention to the definite interconnection between the terms of aqueous balance, which they expressed in the following formulas:

$$E = r \left(1 - e^{-\frac{E_0}{r}} \right) \tag{6}$$

(Schreiber's equation, defined further by Ol'dekop, where E_0 is the maximum amount of evaporation for given conditions) and

$$E = E_0 \cdot \text{th} \frac{r}{E_0} \qquad \left[\text{note: "th" : tanh} \right] \tag{7}$$

(Ol'dekop's equation)

It is easy to prove that both of these formulas satisfy conditions (3) and (4), if we assume that the maximum possible evaporation is equal to the ratio of the heat input by radiation to the latent heat of vaporization, i.e., $E_0 = \frac{R}{L}$.

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In order to verify the above-stated considerations regarding the nature of the relationship (5) for large and small values of $\frac{R}{Lr}$, we shall study the empirical data. From the presently avaliable more or less reliable data regarding the values of the radiation balance, evaporation and precipitation, we shall first note the data obtained by this author [2], T. G. Berlyand [1] and Albrecht [11] at seven different locations (over annual intervals), which embrace a variety of climatic conditions (table 1).

Table 1

The Annual Totals of the Terms of the Thermal and Aqueous Balances

Location	R Kilocalories per square centimeter per year	r Centimeters year	E Centimeters water year	$\frac{E}{r}$	$\frac{R}{Lr}$
Sodankyla (Lapland)	11.6	49	16	0.33	0.39
Potsdam (Mid-Europe)	19.8	58	33	0.57	0.57
Batavia (Java)	56.2	159	74	0.47	0.57
Arlington (USA)	58.0	99	63	0.64	0.98
Irkutsk (Asiatic USSR)	19.9	27	19	0.70	1.22
Tsangyung (Gobi)	42.4	24	22	0.92	2.94
Isma'iliya (Egypt)	36.0	5	5	1.00	12.0

The value of radiation balance for Arlington for the year 1939 obtained in the study [2] is given here after a more accurate determination.

From table 1 it follows that for the comparatively small values of $\frac{R}{Lr}$, in the order of 0.4 - 0.6 (Sodankyla, Potsdam, Batavia), the $\frac{E}{r}$ values approach the $\frac{R}{Lr}$ values, while for the large values of $\frac{R}{Lr}$ (Arlington, Irkutsk) there is observed a shift from the first relationship to the second.

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Figure 1

1 - Lapland; 2 - Central Germany; 3 - Java; 4 - US, Atlantic seaboard,
5 - Irkutsk; 6 - Cold desert; 7 - Egypt

In figure 1 the values of $\frac{E}{F}$ are shown for seven locations and plotted (by means of dots) as a function of $\frac{R}{Lr}$. This graph also indicates the conditions which determine the relationship between $\frac{E}{F}$ and both small and large values of $\frac{R}{Lr}$ (straight line segments 0A and AB).

It must be acknowledged that the location of the points on the graph is in good correspondence with the conditions (3) and (4) and confirms the assumptions made regarding the nature of relationship (5). This agreement is of special significance in view of diversity of the physico-geographic conditions as shown by the experimental data (from desert to subarctic regions). In order to represent the relationship (5) in analytical form, we can either utilize formulas analogous to (6) and (7), such as

$$ E = 1 - e^{-\frac{R}{Lr}} \qquad E = F\left(1 - e^{-\frac{R}{Lr}}\right) \qquad (8) $$

$$ E = \frac{E}{F} \operatorname{th} \frac{R}{Lr} $$

and

$$ E = \frac{E}{F} \cdot \operatorname{th} \frac{Lr}{R} \qquad (9) $$

(equation (8) is shown on the graph by curve I, while equation (9) is represented by curve II), or we can take the geometric mean of the above relationships, which results in: [note: "ch" = cosh, "sh" = sinh]

$$ (10) $$

$$ E = \sqrt{\frac{E}{F} \cdot \operatorname{th} \frac{Lr}{R}\left(1 - \operatorname{ch} \frac{R}{Lr}\right) \operatorname{sh} \frac{R}{Lr}} \qquad E = \sqrt{\frac{E}{F} \operatorname{th} \frac{Lr}{R}\left(1 - \operatorname{ch} \frac{R}{Lr} + \operatorname{sh} \frac{R}{Lr}\right)} $$

which, in turn, is shown on the graph by curve III, the latter being a better fit of the experimental points than the former two curves. Let us note that the choice of any of the cited interpolation formulas does not affect greatly the accuracy of $\frac{E}{F}$ for given values of $\frac{R}{Lr}$. For

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comparatively small and large values of $\frac{R}{Lr}$, the formulas (8), (9) and (10) give practically identical results, while for average values of $\frac{R}{Lr}$ the discrepancies from the average between the values of $\frac{E}{r}$ as determined from formulas (8) and (9) are in the order of 10 percent, which is insignificant compared with the accuracy of determination of the basic parameters.

Another verification of the "connecting equation" can be made according to the data of our determinations over a period of years of the average annual totals of the surface radiation balance of the southern part of the European USSR [8]. For this purpose, there were determined for a number of uniformly spaced locations of a given territory, the average amounts of precipitation in accordance with the GGO chart for 1939, and the average drainage totals in accordance with the chart of B. D. Zaykov [9]. Table 2 gives the results of determinations of the $\frac{E}{r}$ and $\frac{R}{Lr}$ values in accordance with this data, for some 25 locations representing quite different climatic and hydrologic conditions (from Cur yeo and Astrakhan to Moscow and Smolensk).

In figure 2 the graph of $\frac{E}{r}$ versus $\frac{R}{Lr}$ computed by means of the equation (10) is compared with the data of table 2 (dots) and table 1 (circles). It is noted that the distribution of the experimental data confirms rather well the form of the equation which connects the terms of the thermal and aqueous balances at the earth's surface.

This proves, in particular, the possibility of extensive usage of the "connecting equation" in hydrological determination of drainage by indirect methods (without employing the data of hydrometric observations), as well as in the determination of precipitation totals when there is available reliable drainage data.

A detailed study of these problems is intended to be included in a separate paper by this author.

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The Average Annual Totals of the Terms of the Thermal and Aqueous Balances

over a Period of Years

Location	R	r	f	$\frac{E}{R}\cdot 1\cdot\frac{f}{2}$	$\frac{N}{Lr}$
	Kilocalories per square centimeter per year	Centimeters per year	Centimeters per year		
Dir'ev	43	17	5	1.01	4.42
Astrakhan	44	26	0	1.06	4.07
Stalingrad	41	32	4	0.83	2.33
Ural'sk	36	30	4	0.87	2.30
Odessa	43	39	3	0.87	1.80
Kishinev	42	45	7	0.82	1.55
Saratov	35	38	4	0.86	1.54
Voroshilovgrad	41	45	6	0.80	1.5
Dnepropetrovsk	41	49	7	0.74	1.42
Kuybyshev	37	41	6	0.85	1.45
Kharkov	38	52	9	0.87	1.22
Tambov	34	49	12	0.76	1.02
Voronezh	34	52	11	0.74	1.22
Chkalovsk	35	54	12	0.80	1.08
Kiev	36	56	8	0.84	1.02
Kazan	28	47	15	0.68	0.93
Tula	30	56	16	0.71	0.89
Minsk	31	62	24	0.61	0.83
Moscow	28	60	20	0.69	0.78
Smolensk	29	65	22	0.64	0.74

In conclusion let us state that the immediate application of equa-
tions (10), (8) and (9) is possible principally in the case of flat and

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lowland reservoirs, in the case of mountains a change in conditions of
the slope drainage can result in a distortion in the form of the "con-
necting equation," especially in the region of average values of the
$\frac{R}{Lr}$ parameter.

2. On the basis of the above we conclude that the conditions of
thermal and moisture exchange at the level of the underlying surface is
characterized by a system of 3 equations: thermal and aqueous balances
and the "connecting equation." We write these equations in the form of
3 relationships

$$\frac{R}{Lr} = \frac{E}{r} + \frac{P}{Lr}$$

$$1 = \frac{X}{r} + \frac{f}{r} \qquad (11)$$

$$\frac{E}{r} = \phi\left(\frac{R}{Lr}\right)$$

which contain 4 variable values of the relative magnitudes of the terms
of the thermal and aqueous balances.

It is obvious that only one of the variable values in system (11)
is an independent one in view of which we can, for a given value of the
independent variable, determine the relative magnitudes of other terms
of the thermal and aqueous balances. We note, however, that, in view
of the special nature of relationship (5) not any of the variables of
system (11) can be taken as the determining parameter. From the gen-
eral shape of curves shown in figure 1 and 2 it follows that for given
values of $\frac{R}{Lr}$ of certain accuracy, we can always find a value of $\frac{E}{r}$ of the
same, or greater order of accuracy. The reverse problem of determining
$\frac{R}{Lr}$ from a given $\frac{E}{r}$ can be solved sufficiently accurately only for com-
paratively small values of $\frac{E}{r}$, while for $\frac{E}{r}$ values close to unity even
small errors in the determination of this value will lead to large er-
rors in the $\frac{R}{Lr}$ values. Since the relationship (5) is of quasi-empirical

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nature and contains a certain error, then it is evident that, in the region of large $\frac{E}{r}$ values, determination of $\frac{R}{Lr}$ from $\frac{E}{r}$ is practically impossible even with the utmost accuracy of the E and r values. From this, it follows that either $\frac{R}{Lr}$ (or $\frac{P}{Lr}$) must be taken as the determining parameter of system (11), while the variables ($\frac{E}{r}$ and $\frac{c}{r}$) can be considered as determining only for comparatively small values of $\frac{E}{r}$ i.e., for the conditions of moist and cool climates.

Thus, it is established that the relative average values of the terms of the thermal and aqueous balances in a given locality are defined by the numerical value of a single parameter ($\frac{R}{Lr}$, for example). The absolute values of the terms of the thermal and aqueous balances are determined by two parameters (in this case we have 5 variables and 3 equations) such as, for example, R and r, or r and $\frac{R}{Lr}$ (here also, in view of the nature of the relationship (5), such parameters as, for example, E and r, etc., cannot be considered as determining for all climatic conditions.

In accordance with the above considerations let us assume that the numerical values of parameters, which determine the terms of the thermal and aqueous balances of the underlying surface, characterize the climatic factor of the surface physico-geographical process, which, in turn, determines development of vegetation, river flow and soil origins.

In order to check this hypotheses we compare the distribution of the $\frac{R}{Lr}$ isolines with the boundaries of the basic physico-geographical zones; let us observe that the difficulty of this computation is due to the limited amount of existing data in connection with the distribution of average annual totals of radiation balance at the earth's surface compiled over a period of years. (The use of the other determining parameter $\frac{P}{Lr}$ does not simplify the problem at hand, since computation of the geographical distribution of P is no less complex than that of geographical distribution of R.)

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In particular, it is pointed out that the data on radiation balance of the several locations in table 1 is useless in this case, since, besides being very scant, this data refers to isolated years and small areas of observations, not always representative of the surrounding territories. In order to determine the distribution of the parameter $\frac{B}{L_r}$, we use, before everything else the chart of average totals of radiation balance compiled over many years for the southern portion of the European USSR $\sqrt{3\_7}$. This chart can be supplemented for the northern part of the European USSR by using the equation (10) in determining the radiation balance, since for this region there is available comparatively exact precipitation (a chart by O. A. Drozdov) and drainage (a chart by B. D. Zaykov) data, where equation (10) is sufficiently accurate, since the value of $\frac{F}{\mu}$ does not exceed 0.7.

This chart of the surface radiation balance of the European USSR is shown in figure 3. (Isolines south of the 57 degree Northern latitude were determined by means of actinometric observations and computations $\sqrt{3\_7}$; those north of the 55 degree latitude were determined by means of equation (10), the values at isolines being given in kilocalories per square centimeter per year.)

Examination of this chart leads to the conclusion that the average annual totals of radiation balance over flat earth surfaces in moderate latitudes are basically latitudinal characteristics. In paper $\sqrt{3\_7}$ it is noted that this phenomenon is determined by the mutual compensation in the action of a number of factors, which vary materially in accordance with latitude. Thus, to a first approximation, we can consider that, over flat earth surfaces in moderate latitudes, the distribution of the radiation balance can be characterized by the average dependence of R on latitude. (Figure 3. Chart of the Radiation Balance over the Surface of the European USSR.)

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The first determination of the mean latitudinal distribution of
radiation balance over the earth's surface was made by this author and
M. I. Yudin [4] in a computation, which, however, was somewhat sche-
matic in view of the utilization of rather inaccurate radiation and emis-
sion data of N. C. Yevfimov. For a more precise determination of the av-
erage relationship between radiation balance and latitude, we can use
the data of figure 3, taking into account the results of other compu-
tations performed by the Main geophysical observatory.

In figure 4 the average latitudinal distribution of the radia-
tion balance over the earth's surface obtained from the data of figure 3
is shown by the curve A. Curve B shows the average latitudinal distri
bution of the radiation balance for the European USSR as obtained by T. G.
Berlynd [1] entirely by means of actinometric data. Curve C shows
the results of computations by L. I. Zubenok, who computed the average
latitudinal distribution of annual totals of thermal losses F over the
European USSR, on the basis of network gradient observations, and de-
termined the radiation balance b summing F and LE (evaporation was de-
termined by the precipitation and drainage data)

It is seen that the results of these 3 independent computations
show good agreement in the region of 50 to 60 degrees of Northern lati-
tude. For latitudes of 45 to 50 degrees this author's and Berlyand's com-
putations show greater values of the radiation balance than shown by
Zubenok's computation. North of 60 degrees the values of the radiation
balance obtained by Berlyand are greater than those obtained by this
author, Zubenok's data for this region being intermediate of these two
computations. At the extreme northern portion of the European USSR
(68 - 70 degrees) the average values of radiation balance as obtained by
this author and Berlyand approximately coincide. On the basis of re-
sults shown in figure 4, and taking into account their accuracy for vari-
ous latitudes, as well as in accordance with the radiation balance data

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from equation (10), we have adopted as the mean latitudinal distribution of radiation balance over the flat earth regions the curve in figure 5, which approaches closely curve A of figure 4.

The most reliable distribution of the mean annual precipitation totals over the USSR territory is given by Drosdov's chart, which, like other precipitation charts, cannot be considered as sufficiently accurate for Eastern Siberia. By means of the Drosdov chart and curve of figure 5 distribution of the $\frac{R}{Lr}$ parameter over the European USSR was found, as well as over Western Siberia and Central Asia (USSR territory west of 90 degrees Eastern longitude). Thus obtained isolines $\frac{R}{Lr}$ corresponding to the values 0.35 (1); 1.1 (2); 2.3 (3); 3.4 (4) are shown in figure 6. Comparing this chart with the chart of figure 7 which shows boundary lines between the vegetations of tundra, forest zone, steppes, semidesert and desert zones it is easily seen that these isolines define very accurately the boundary lines of the basic phyto-geographical zones. (The chart of figure 7 was drawn up in accordance with the GGO charts, 1937 (which are somewhat schematic))

An analogous computation was made for the North American territory, where it was established that the boundaries of the phytogeographical zones are defined rather well by the same values of the $\frac{R}{Lr}$ parameters (high mountain regions are shown cross-hatched). On the basis of these data we can conclude that the parameter $\frac{R}{Lr}$ is actually the determining characteristic of the biological (phytogeographical) factor of the surface physico-geographical process. Since the geographical distribution of soils is closely connected with the distribution of vegetation, the determining significance of the parameter $\frac{R}{Lr}$ in the process of soil formation must be recognized.

(Figure 7. Boundary lines between the basic phyto-geographical zones (high mountain regions shown as cross-hatched).)

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Dependence of the hydrological factor of the surface physico-geographical process on the characteristic of $\frac{R}{Lr}$ is the immediate consequence of system (11); whence it follows that the coefficient of drainage is determined, by a single meaning value of $\frac{R}{Lr}$. In this connection it must be noted that to each phytogeographic zone there must correspond a definite range of values of drainage coefficients, the limits of which can be found from the curve of figure 1 and the above-cited values of $\frac{R}{Lr}$ at the boundaries of the phytogeographic zones

According to the data of such calculation we find that on the tundra forest boundary the drainage coefficient is approximately equal to 0.7; at the forest-steppe boundary it is equal to 0.2-0.3; at the steppe semidesert boundary it is equal to 0.05-0.1 and at the semidesert desert boundary it is in the neighborhood of several hundredths. It should be noted that these results are in good agreement with the data obtained by C. K. Davydov [2] for the European USSR, who, by the analysis of empirical data, found that in the tundra zone the drainage coefficient reaches the value of 0.8 and larger, while in the forest zone it decreases gradually to 0.2-0.3 and is of the order of 0.08-0.22 in the steppe regions.

Thus, it can be stated that the $\frac{R}{Lr}$ parameter, according to the above considerations, determines the zonality of distribution of geological regimes, vegetation and soil species. We note that in developing some of the assumptions on paper [3] it is easy to establish the connection between the characteristic of $\frac{R}{Lr}$ and the conditions of intensity of the surface physico-geographical process.

The above data indicate that for small values of the $\frac{R}{Lr}$ parameter there are created the conditions for excessive moisture, and for large values of $\frac{R}{Lr}$ there are created the conditions for insufficient moisture

In accordance with Grigor'yev's conception it can be concluded that, for both of these extreme instances, the intensity of the surface physico-geographical process decreases as compared with the conditions for optimum moisture, observed for average value of $\frac{R}{Lr}$. Since the

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$\frac{R}{Lr}$ parameter determines the magnitude of the drainage coefficient, it also regulates the conditions of the soil salt balance; from this viewpoint small and large values of $\frac{R}{Lr}$ also decrease the intensity of the surface physico-geographical process, since in the first case large drainage results in the removal of carbonates, while in the second case small drainage might result in conditions favorable for excessive salt accumulation in the soil. In view of this the $\frac{R}{Lr}$ parameter becomes the sole characteristic which determines the intensity conditions of the surface physico-geographical process. Taking into account the optimum conditions for moisture and salt balance, it is possible to evaluate approximately $\frac{R}{Lr}$ corresponding to the maximum intensity of the surface physico-geographical process, which, evidently is in the order of 1.2. In other words it can be shown that the greatest intensity of the surface physico-geographical process for a given "energy basis" (determined by the value of R) is observed when the quantity of heat necessary to evaporate completely of the annual precipitation totals comprises 80-100 percent of the magnitude of the radiation balance. For relatively larger or smaller precipitation totals, the intensity of the surface physico-geographical process is decreased compared with the maximum possible intensity.

The above-cited considerations regarding the determining parameters of the terms comprising the equations of the thermal and aqueous balances at the earth's surface can be also used in working out the classification of climatic zones. This type of classification is along the trend marked out in the known works of Köppen, Voyeykov and Penk, who analyzed climate zonality according to the characteristic influence of climatic factors upon a given natural process (vegetable life or a hydrological regime). As compared with the schemes of these authors, the above suggested classification has two advantages:

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1. Boundaries of the climatic zones in our classification will at the same time determine the zonality in the distribution of all the other basic factors of the surface physico-geographical process (river flow, vegetation, soil formation).

2. Boundaries of the climatic zones will be determined according to the distribution of parameters, which have a definite physical meaning, instead of on the basis of purely empirical characteristics.

From general considerations it is obvious that in order to work out a sufficiently detailed climatic classification we must use characteristics which determine the absolute values of the terms of equations of thermal and aqueous balances; i e., two determining parameters (for example, r and $\frac{r}{Lr}$).

Classification based on this principle will be primarily of physico-geographical importance, as distinguished from purely meteorological classifications, which use the internal characteristics of the atmospheric processes.

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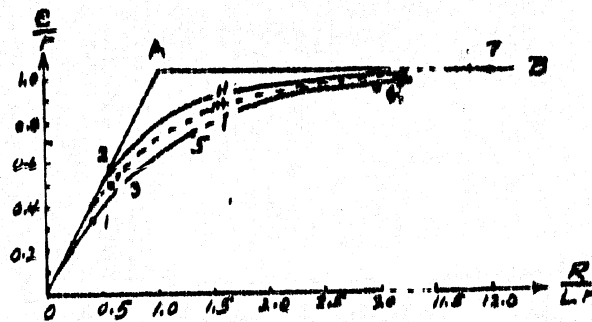
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Figure 1.   1: Leningrad; 2: Middle Germany; 3: Yava;
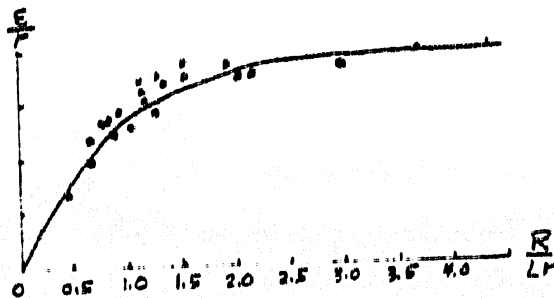4: USA, Atlantic Coast;   5: Irkutsk ; 6: Gobi; 7: Egypt.

Figure 2.

Figure 4.

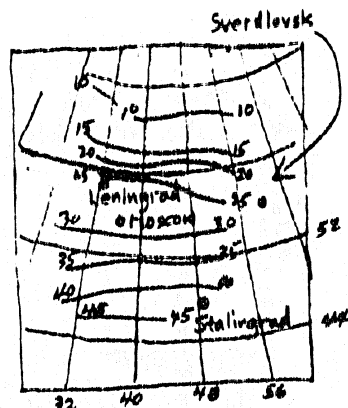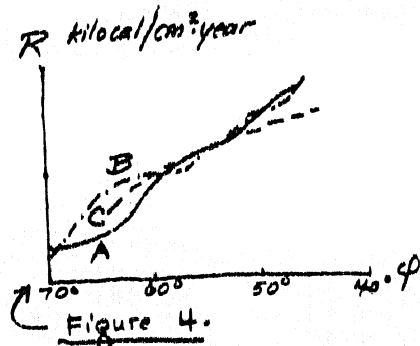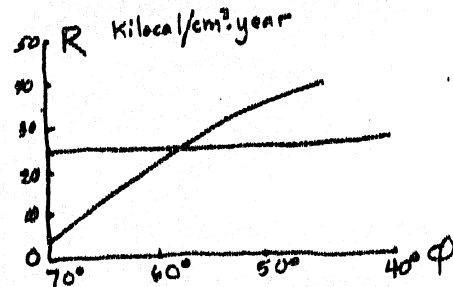Figure 3. Map of Radiational
Balance of the Surface.

Figure 5.

-A-
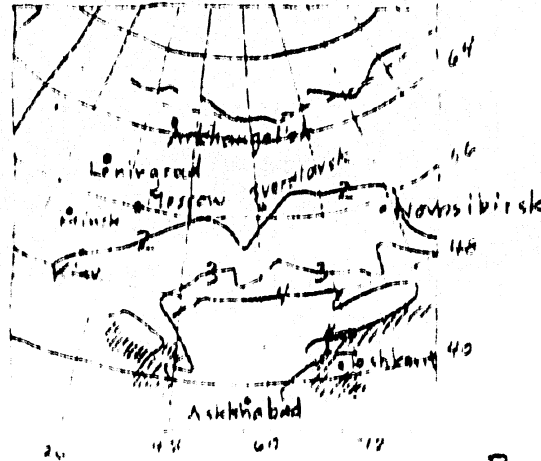
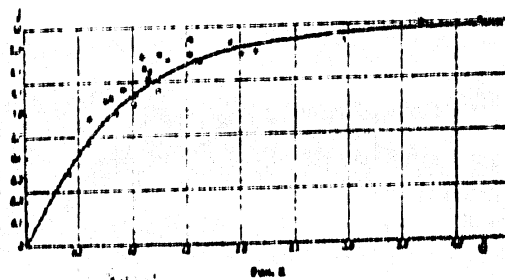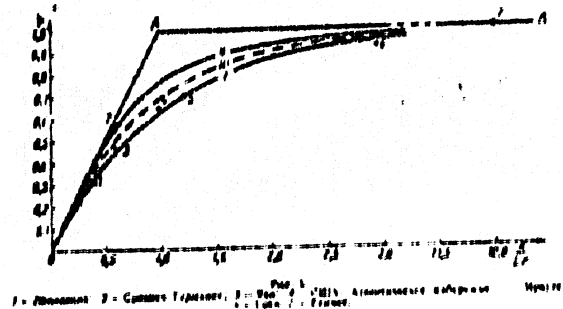Figure 6. Isolines of the Parameter $\frac{R}{L_r}$ (high-mountainous regions are shaded).
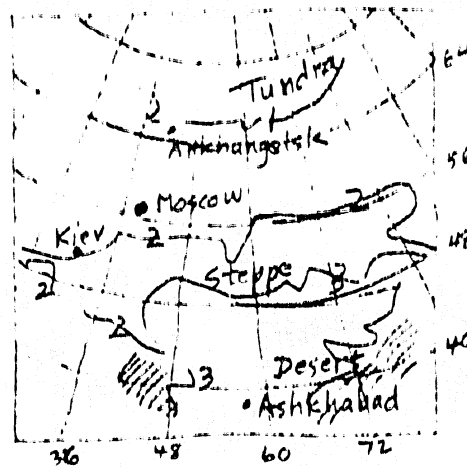
Figure 7. Boundaries of Main Phyto-geographic Zones (High-Montainous Regions are shaded).

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[These are photostats of the original figures
1 and 3, that happen to be available]